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PREDICTION OF THE DOSAGE TO MAN FROM THE FALLOUT OF NUCLEAR DEVICES
6. TRANSPORT OF NUCLEAR DEBRIS BY SURFACE AND GROUNDWATER

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Abstract

This report presents a two-part discussion of the surface and groundwater transport of the radionuclides and debris produced by a large-scale nuclear cratering detonation such as might be used in the course of a nuclear excavation project.

The first section briefly discusses the transfer of crater debris to surface-water systems associated with the excavation site as the result of erosion and the subsequent transport of debris, through the combined actions of rainfall and runoff.

It makes two major points. First, the amount of debris that can be eroded as the result of a single storm, or a series of storms, is a function of (1) the amount of rainfall, its intensity, and the distribution of intensity with time, (2) the infiltration and erosibility characteristics of the debris, (3) the topography, length, slope, and size of the area being eroded, and (4) the type and nature of vegetative cover, if any. Second, the amount of debris that can actually be transported to the surface water system of concern is governed by (1) the amount of debris eroded, (2) the amount, velocity and turbulence of the run-off water, and (3) the distance from the erosion site to the stream channel. Since most of these variables vary widely from area to area, each proposed excavation site presents a unique problem that requires hydrologic and topographic study specifically directed to the determination of the amount of crater debris that can be transferred to an aquatic system associated with that site. In the absence of such detailed knowledge, estimates of the potential debris transfer can be made by estimating the amount of runoff and the debris transfer coefficient. This procedure shows (1) that in regions of little or no rainfall, there will be negligible movement of debris towards aquatic systems, except by atmospheric transport, (2) that a single intense storm that produces 3 inches of runoff may cause as much as 1% of the total debris to reach a nearby stream channel, and (3) that in areas of heavy rainfall, such as the Panama Canal Zone, about 20% of the total debris might reach an aquatic system in a year's time.

The second section of the report briefly examines the Darcy's Law description of groundwater flow and shows

that in certain geologic situations, the groundwater travel-times and paths predicted on the basis of standard field measurements of aquifer properties, especially the permeability, can be highly misleading with regard to the actual movement of a significant portion of the contaminated water. The reason for this is that field measurements determine overall average values of aquifer properties in accordance with two assumptions: (1) that water-bearing formations are both homogeneous and isotropic, and (2) that groundwater flows in response to the average hydrologic properties of this idealized formation. However, real geologic strata exhibit great variations in properties, particularly in permeability with depth and direction, as well as large-scale inhomogeneities such as faults and channels. (As used in this report, "inhomogeneous" refers to a more or less continuous variation in permeability; "heterogeneous" refers to a discontinuous change in permeability.) Consequently, unless measurements are sufficiently detailed to resolve the effects of these deviations from ideal behavior, contaminated water may either (1) arrive where it is expected, but much sooner than expected, perhaps by two orders of magnitude, (2) arrive where it is not expected at all, or (3) spread and travel in a much broader band than expected, with the concentration front arriving considerably before the peak concentration.

Introduction

Besides the direct deposition of atmospheric fallout, there are two main modes by which substantial amounts of radioactivity can be introduced into the aquatic environment during the course of a nuclear excavation project:

- Erosion and subsequent transport of crater debris to surface water systems associated with the crater site by rainfall and runoff.
- Direct injection of radionuclides into, and subsequent transport through, the excavation site groundwater system.

Thus, before a nuclear excavation project is executed, the potential dosage to man from each of these processes should be estimated to ensure that unacceptable levels of radiation could not arise from the project. Such an evaluation can be made using the procedures detailed in UCRL-50163, Part 5 (Ref. 1), provided that it is possible to determine the fraction of the radioactive material in equilibrium with the aquatic system.

For the case of surface water, we assume that whatever debris reaches the aquatic system equilibrates with that system. Thus, the problem is to determine what fraction of the total crater debris reaches that system. For the case of groundwater, we assume that the various radionuclides are distributed in accordance with the amount of ion exchanger they encounter. Thus the problem becomes one of determining the aquatic half-life of each nuclide, i.e., the time required for the volume of water

containing this nuclide to be displaced from the point of injection to the point of concern.

Thus, the first section of this report is concerned with the transfer of crater debris to aquatic systems as the result of erosion and subsequent transport, through the combined actions of rainfall and runoff. The second section discusses some of the difficulties involved in predicting the groundwater travel-time of radionuclides on the basis of standard field measurements of aquifer properties.

Surface-Water Transport of Crater Debris

This section of the paper discusses the potential transfer of crater debris to an aquatic system as the result of erosion and subsequent transport through the combined actions of rainfall and surface run-off. It shows that the amount of debris transferred as the result of a single storm or series of storms is a function of the intensity of the rainfall, the distribution of its intensity with time, and the duration of rainfall; the infiltration and erosibility characteristics of the debris; the capacity of the runoff water to transport debris and the topography of the crater site; and the dimensions of the area over which the debris is dispersed. Thus at any particular site, prediction of the amount of debris that can be transferred to an aquatic system requires a detailed study of the hydrology and topography of that site.

The transfer of debris to an aquatic system through the combined action of rainfall and runoff is a twofold process in which rainfall is responsible for the initial erosion (or detachment) of debris from the land surface through the energy of raindrop impact, and surface runoff is responsible for the subsequent downslope transport of this eroded debris towards stream channels.[2,3,4] To a good approximation, the amount of runoff water available to perform this function is the difference between the total rainfall and the amount of water that enters the soil during the course of the storm. Thus, if there is little or no runoff, there will be little or no movement of eroded debris toward collecting streams.

In what follows we discuss the relationships among rainfall, runoff, and soil loss in order to determine some of the geologic, hydrologic, and topographic factors that govern the transfer of crater debris to surface water systems. Finally, we define a parameter that allows us to estimate the maximum amount of debris that can undergo transfer in various situations.

INFILTRATION

During a rainstorm, water enters the soil by a process known as infiltration. The maximum rate at which a soil can absorb water is called its infiltration capacity. The factors that influence infiltration,[5,6] are:

(1) Moisture content of the soil.

(2) Clay content of the soil.

(3) Amount and intensity of the rain.

(4) Physical properties of the soil (porosity, particle-size distribution, cohesion, condition of soil surface).

(5) Amount and type of vegetation.

In general, all these factors interact in such a way that infiltration has a high initial rate that diminishes to a nearly constant rate during continued rainfall. This relationship can be expressed by an equation of the form[6]

$$f_t = f_c + (f_0 - f_c)e^{-kt}, \qquad (1)$$

where

f_t is the infiltration rate (in./hr) at any time t,

f_0 is the infiltration capacity (in./hr) at the beginning of the storm,

f_c is the final constant rate of infiltration (in./hr),

k is a constant for a particular soil and vegetative cover; the normal range of values is from 2.0 to 10.0 hr^{-1}.

The total amount of water that enters the soil (i.e., the total infiltration F in inches) during a rainstorm is given by the integral of the infiltration rate over the length of the storm:

$$F = \int_0^t f_t \, dt = \int_0^t \left[f_c + (f_0 - f_c)e^{-kt}\right] dt, \qquad (2)$$

where t is the duration of the storm in hours. For most values of f_0, f_c, and k, the total infiltration F at any time t can be approximated within a factor of two or better by the expression

$$F \approx f_c t \, . \qquad (3)$$

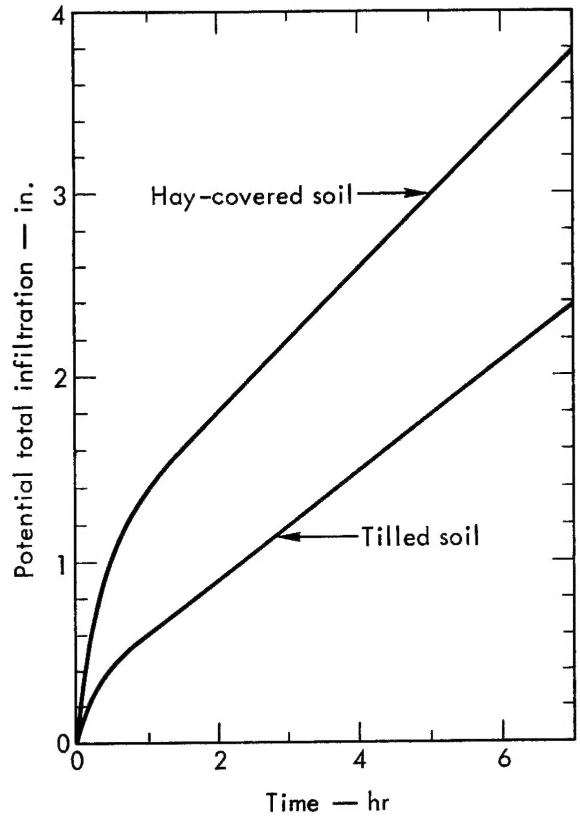

Fig. 1. Potential total infiltration as a function of time for tilled and hay-covered soils.

For purposes of this paper, a crater site is regarded either as barren soil or as planted with a protective ground cover. It will be assumed that in the first case the soil has infiltration characteristics comparable to those of tilled soil, whereas the covered soil has characteristics comparable to those of a hayfield. Total infiltration curves for tilled and covered soils as a function of time (Fig. 1) were prepared based on the data of Holtan and Kirkpatrick.[7]

RUNOFF

When the rainfall rate exceeds the infiltration capacity of the soil, a thin sheet of water builds up and begins to flow over the land surface. On its way

Table 1. Potential run-off from tilled or hay-covered soil from storms of various intensities.[a]

Place	Date	Intensity (in./hr)	Duration (hr)	Run-off (in.) Tilled soil	Run-off (in.) Hay-covered soil
Opids Camp, Calif.	Aug. 1926	61.8	0.017	1.0	0.9
St. Louis, Mo.	Aug. 1848	20.2	0.25	4.8	4.3
Guinea, Va.	Aug. 1906	18.5	0.50	8.8	8.1
Galveston, Tex.	June 1871	17.0	0.23	3.7	3.2
Buffalo, N.Y.	Mar. 1897	15.8	0.05	0.7	0.6
Augusta, Ga.	June 1911	14.9	0.083	1.1	0.8
Norfolk, Va.	Aug. 1888	14.9	0.17	2.3	2.0
Taylor, Tex.	Sept. 1921	12.0	0.67	7.5	6.8
Brownsville, Tex.	Oct. 1884	11.2	0.10	1.1	0.8
Catskill, N.Y.	July 1819	10.0	1.0	9.4	8.6
Sandusky, Ohio	July 1879	9.0	0.25	2.0	1.5
Erie, Pa.	June 1886	8.1	0.25	1.7	1.3
D'Hanis, Tex.	May 1935	7.3	3.0	20.6	19.6
Springbrook, Mont.	June 1921	6.4	1.08	6.3	5.4
Catskill, N.Y.	July 1819	2.4	7.5	15.4	14.0
Gatun Dam, Panama	Dec. 1939	1.4	6.0	6.2	4.9
Hearne, Tex.	June 1899	1.0	24.0	16.5	13.4
Kaplan, La.	Aug. 1940	0.96	24.0	15.5	12.4
Taylor, Tex.	Sept. 1921	0.96	24.0	15.6	12.5
Alta Pass, N.C.	July 1916	0.93	24.0	14.7	11.6
Springbrook, Mont.	June 1921	0.82	14.0	7.0	4.9
Aqua Clara, Panama	Dec. 1944	0.57	24.0	6.0	3.0
Beaulieu, Minn.	July 1909	0.45	24.0	3.2	0.2

[a]U.S. storm data from Bernard,[8] Panama data from Long Range Canal Report.[9]

to an established stream channel this flow is known as overland flow; once it reaches the channel, it is known as surface runoff.[2]

To a good approximation, the amount of runoff water available due to a storm is the difference between the total rainfall and the total infiltration during the storm, where the infiltration is given by Eq. (2) or approximated by $f_c t$. From Fig. 1 it can be seen that runoff will not be significant on either tilled or hay-covered soil unless rainfall intensity exceeds 0.5 in./hr. Much more intense storms than this have been observed; a selected list of these together with the associated values for potential runoff from tilled and hay-covered soil is presented in Table 1. If the soil layer immediately below the debris layer has significantly smaller infiltration characteristics than those postulated above,

however, the infiltration rate will be lower. As a result, the runoff would be greater than would be predicted on the basis of the above characteristics. Thus, in certain cases it would not be unreasonable to assume that all of the rainfall becomes runoff.

SOIL EROSION

The amount of soil that can be eroded from an area subject to raindrop impact as the result of a single storm or a series of storms can be estimated from a general equation based on the work of Wischmeier and his associates[10-12]:

$$S \approx 8 \times 10^{-2} E_R S_E I_{max}^{30} P_C L_S, \quad (4)$$

where

S is the soil loss (g/ft^2) due to a single storm,

E_R the relative soil erosiveness factor, is the tendency of a particular soil to be detached and carried away (it is a function of the proportions of sand, silt, clay, colloidal material, and organic matter comprising the soil),

S_E the storm energy, is the summation over the intensity distribution of the kinetic energy at each intensity times the amount of rainfall that falls at that intensity (the kinetic energy at any particular intensity can be read from Fig. 2, which is based upon the relation[10] $KE = 916 + 331 \log_{10} I$, where I is the intensity in inches per hour and KE is the kinetic energy in foot-tons per acre-in.),

I_{max}^{30} is the maximum 30-min rainfall intensity (in./hr) (for storms of less than 30 minutes duration, I_{max}^{30} should be replaced by the average intensity; the product of I_{max}^{30} and the storm energy is a good measure of the capacity of the storm to erode soil from an unprotected field[10,11]),

P_C is the plant coverage factor, defined to be unity for bare or fallow soil,

L_S the land slope factor, is the ratio of the product of length

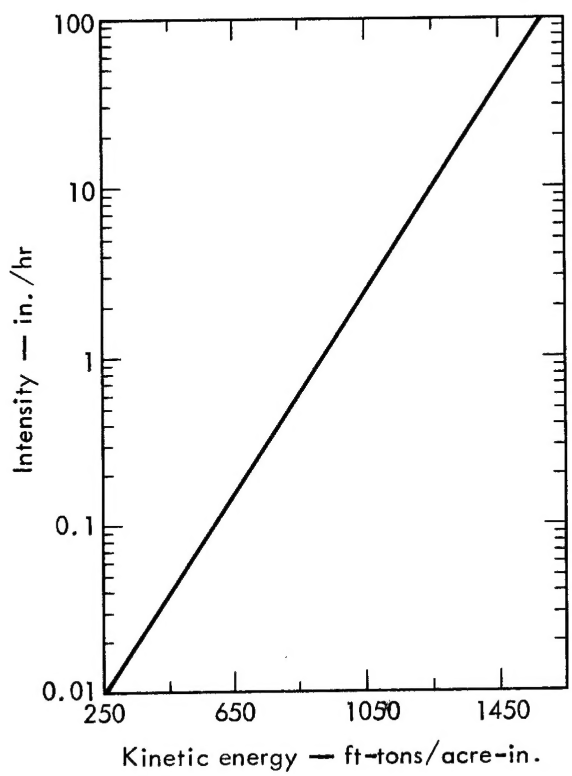

Fig. 2. Kinetic energy of rainfall as a function of intensity.

and slope for the area being eroded to that for a standard plot of 9 deg slope and 73-ft length, i.e., 657 deg-ft (data developed by the original authors and others show that, in general, this equation is quite accurate and capable of predicting between 70% and 120% of the observed soil loss[10,11,13]).

Measurements from standard plots[12,14] indicate that most soils can be grouped as shown in Table 2 with regard to relative erosibility. These data indicate that the relative erosibility of most soils varies only by a factor of 2 on either side of that for the standard prairie soil; the variation is probably due to differences in permeability and/or texture of the various soils. In the absence of actual information about the relative erosibility of cratering debris or sufficient data about its physical properties to infer its relative erosibility, a conservative value of $E_R = 2$ could be used to estimate the erosion of debris.

Table 2. Relative erosibility of various soil types.

Type	Relative erosibility
Silty loams	0.50 - 1.25
Prairie soils	1.00
Soils developed under forest cover	1.25
Slowly permeable soils and some sandy soils	1.50
Soils with tight, nearly impermeable sub-soils	1.75

DEBRIS TRANSFER COEFFICIENT

On the basis of the information presented in the previous sections, the amount of debris that can be eroded as the result of a single storm or a series of storms is a function of (1) the characteristics (intensity, distribution of intensity with time, and duration) of the storm, (2) the erosibility of the debris, (3) the length, slope, and size of the area being eroded, and (4) the type and amount of vegetative cover, if any.

The ability of the runoff water to transport this eroded material to a surface water system is a function of the amount of runoff water, its velocity and turbulence, and the distance to the stream channel. The amount of runoff is a function of the intensity and duration of the rainfall and the infiltration characteristics of the debris. Velocity of runoff is primarily a function of the slope, nature, and topography of the land surface over which it flows. Turbulence of runoff is a function of velocity and raindrop impact.

From the above, it can be seen that the major site-dependent variables involved in determining the amount of debris that can be transferred to an aquatic system are the characteristics of the rainfall available to erode the debris, and the topography, particularly the distance to stream channels, over which runoff must transport the debris. It would, therefore, be useful to find a parameter that would combine these factors and allow one to estimate the maximum potential transfer of debris in a simple fashion. Since the potential transfer of debris is limited both by the

Table 3. Soil loss/runoff weight ratios observed in various situations.

Description of plot	Rainfall characteristics	Wt. soil loss / Wt. of runoff	Ref.
12 ft long, 36 ft wide, soil taken from timothy meadow	Artificial rain, I = 4 in./hr, applied for 1 hr in 10-min intervals	0.07-0.18 (16-deg slope)	15
		0.03-0.11 (8-deg slope)	15
12 ft wide, 35 ft long, bare soil, Zanesville silt loam	Artificial rain, intensity of 2.5 in./hr applied as follows: 1 hr, then 24-hr delay; then 1/2 hr, then 15 min delay, then another 1/2 hr	0.05	16
630 ft long, uniform slope, cultivated plot	Rainstorm, $\overline{I}$ = 0.6 in./hr for 5 hr	0.32	17
	Rainstorm, $\overline{I}$ = 1.1 in./hr for 1 hr	0.16	17
	All storms Aug. 1932-Dec. 1934	0.085	17
315 ft long, uniform slope, cultivated plot	Rainstorm $\overline{I}$ = 0.6 in./hr for 5 hr	0.25	17
	Rainstorm, $\overline{I}$ = 1.1 in./hr for 1 hr	0.12	17
	All storms Aug. 1932-Dec. 1934, averaged	0.05	17
157.5 ft long, uniform	Rainstorm, $\overline{I}$ = 0.6 in./hr for 5 hr	0.24	17
	Rainstorm, $\overline{I}$ = 1.1 in./hr for 1 hr	0.08	17
	All storms Aug. 1932-Dec. 1934, averaged	0.025	

amount of runoff water available and by the ability of the water to carry the eroded material from its source area to the stream system of concern, let us therefore define the debris transfer coefficient DTC as the ratio of the weight of the sediment that reaches the stream channel to the weight of the runoff water that transports it there, i.e.,

$$DTC = \frac{\text{weight of sediment}}{\text{weight of runoff}} . \qquad (5)$$

Thus, the weight of debris that is transferred (DT) is

$$DT = DTC \times \rho \times A \times R_u , \qquad (6)$$

where

DTC is the debris transfer coefficient as defined by Eq. (5) above,

ρ is the weight of 1 ft^3 of runoff (2.8×10^4 g/ft^3),

A is the area being eroded by rainfall (ft^2) and

R_u is the runoff due to rainfall (ft).

The question, then, is what is a reasonable value or range of values for DTC? Two types of data bear on this question:

1. Soil-loss data, which define the amount of material detached from

a plot of ground under varying conditions of rainfall, topography and vegetative cover, but give little or no idea of how far it can be transported. These data indicate that for a short transport phase, the DTC often range from 0.05 to 0.25, with an upper limit of 0.3 (Tables 3 and 4) and are consistent with the range and upper limits for watershed[19] and river concentrations of suspended sediments.[20-22]

2. Sediment delivery ratio data measure the ratio between the eroded material that reaches a specific point in a watershed and the total or gross erosion of material above that point. These data[23] indicate that (1) in small watersheds, i.e., those in which the distance to a collecting stream is of the order of 1 mile, about 50%, on the average, of the eroded material will reach the stream system; (2) in medium-sized watersheds, i.e., those in which the distance to a collecting stream is of the order of 3-4 miles, about 25%, on the average, of the eroded material will reach the stream system; (3) in large watersheds, i.e., those in which the distance to a collecting stream is of the order of 5-15 miles, about 15%, on the average, of the eroded material will reach the stream system.

It seems, therefore, that a reasonable upper limit of DTC, which can be used for purposes of estimating debris transfer until more detailed information is made available, can be obtained by multiplying 0.3, which is the maximum ratio of observed soil loss to runoff weight for experimental plots, by 0.5, the sediment delivery ratio for short transport distances,

$$DTC = (0.3)(0.5) = 0.15.$$

In most cases of interest, this value should lead to results that are accurate within a factor of 2.

ILLUSTRATIVE EXAMPLE: THE PANAMA CANAL ZONE

For illustrative purposes let us consider the transfer of debris to an

Table 4. Soil loss/runoff weight ratios observed in Utah on an artificially denuded plot (adapted from Ref. 18)

Storm date	Total rainfall (in.)	Runoff (in.)	Ratio[a] $\frac{\text{wt. soil loss}}{\text{wt. runoff}}$
July 10, 1950	0.7	0.4	0.25
Aug. 19, 1951	1.2	0.6	0.16
Aug. 4, 1954	1.2	0.4	0.10
Aug. 19-20, 1959	1.0	0.4	0.11
Sept. 3, 1960	0.6	0.2	0.25
Aug. 25, 1961	0.6	0.2	0.20
July 13-14, 1962	2.6	1.0	0.18

[a]Estimated on the basis that 1 ft^3 of soil weighs 100 lb.

aquatic system at a 10 Mt crater site like those that would be produced in the proposed excavation of a sea level canal through the Isthmus of Panama.

Observations at Sedan Crater[24,25] indicate that the region of heavy deposition of debris and extensive destruction of vegetation extends outward to a distance of about 5×10^3 ft from ground zero. The relationships between yield and base-surge radius[26] suggest that for a 10-Mt row crater the corresponding distance would be about 2×10^4 ft. Thus the surface area expected to receive heavy deposition of debris encompasses approximately 10^9 ft^2.

In the area under consideration, the infiltration characteristics are similar to those of bare soil for most of the area, falling off somewhat towards the outer boundaries of this area. Thus, for the continuous type of rainfall that occurs in this region,[9,27] it seems reasonable to assume that at least 50% of the rain falling on this area of heavy deposition would become runoff. In a typical year the rainfall averages 120 inches, but as much as 240 inches of rainfall has been recorded.[9] Consequently, about 5 ft of runoff might be expected in a year's time.

From a detailed map of the proposed Route 17 (Sasardi-Morti) area,[28] we see that for many of the shots proposed for this project, the transport distance from the area of heavy debris deposition to a collecting stream will be of the order of 1 mile or less. Thus it seems reasonable to use the estimated upper-limit a value of 0.15 for DTC.

Therefore, from Eq. (6),

$$DT = 0.15 \times 2.8 \times 10^4 \text{ g/ft}^3 \times 10^9 \text{ ft}^2 \times 5 \text{ ft} \approx 2 \times 10^{13} \text{ g}.$$

Since it has been estimated that the total debris associated with such a 10-Mt shot would be of the order of 10^{14} g, the above example thus suggests that, on the average, 20% of this debris could be transferred to the associated aquatic system in a year's time. However, it should once again be pointed out that if the underlying soil layer has significantly smaller infiltration characteristics than the debris layer, the runoff might be a significantly larger fraction of the rainfall than is used in this example. In certain cases, in fact, knowledge of the preshot soil profile might indicate that proper estimate should properly be based on 100% of the rainfall going to runoff.

CONCLUSIONS

This part of the report discusses the processes by which crater debris is transferred by erosion to surface-water systems associated with a nuclear excavation site, and the subsequent transport of the eroded debris through the combined action of rainfall and surface runoff. It is shown that the amount of debris that can be transferred to an aquatic system at any specified site will depend upon the geology, hydrology, and topography of that site in a complex way. Among the variables governing this process are (1) the amount of rainfall, (2) its intensity, (3) the infiltration and erosibility characteristics of the debris, (4) the length,

slope, and size of the area over which the debris is dispersed, (5) the type and nature of vegetative cover, (6) the amount, velocity and turbulence of runoff and (7) the distance from the crater site to a collecting stream.

It is also shown that in the absence of detailed data about a specific excavation site, the potential transfer of debris can be estimated on the basis of knowledge of the potential runoff and its debris transfer coefficient. Application of this procedure shows that although in regions of little or no rainfall, little or no debris would be moved toward stream channels, but in regions of heavy rainfall (e.g. the Panama Canal Zone), between 10 and 40% of the total debris produced by a nuclear cratering detonation could reach an aquatic system within a year's time.

Groundwater Transport of Radionuclides

This part of the paper discusses the time required for the volume of water containing each nuclide to be displaced from the point of injection into the groundwater system to the point of concern. This variable, known as the aquatic half-life, is a function of both the radiological decay rate of the isotope and the rate of mixing with uncontaminated water.[1] For purposes of this presentation we confine ourselves to the case where there is no mixing; thus we are concerned only with determining the radionuclide travel time between the two points in question. Since the various radionuclides for the most part migrate along with the general groundwater flow but travel at slower speeds because of the physiochemical reactions they undergo with the aquifer materials, the maximum rate of transport (minimum travel time) can be determined from a study of the destination and time of arrival of the water pulse.[29,30] Then, the travel time for any specific radionuclide can be determined by multiplying the minimum travel by a suitable factor that accounts for the physiochemical retardation of the nuclide. Such studies are extremely difficult to carry out, however, because of (1) the remoteness of the flow system, (2) the three-dimensional nature of the problem, and (3) the varying characteristics of the geological strata. Fortunately, the problem can be simplified greatly by using tritium as an indicator of the behavior of the groundwater, since this nuclide travels at virtually the same speed as the groundwater pulse.

This part of the paper therefore briefly examines the Darcy's-Law description of groundwater flow and shows that for certain geological situations, estimated tritium travel times and paths based on standard field measurements of aquifer properties, especially permeability, may be highly misleading with regard to the actual movement of a significant portion of the contaminated water. The reason is that field measurements determine overall average values of aquifer properties in accordance with two assumptions: (1) water-bearing formations are both

homogeneous and isotropic, and (2) groundwater flows in response to the average hydrologic properties of the idealized formation, whereas real geological strata not only vary greatly in their properties (particularly permeability) with depth and direction, but also have heterogeneous sections. It will be shown that as a consequence, unless measurements are sufficiently detailed to resolve the effects of these deviations from ideal behavior, tritium-contaminated water either (1) can arrive where it is expected but much sooner than expected, perhaps by two orders of magnitude, (2) can arrive where it is not expected at all, or (3) can spread and travel in a much broader band than expected, with the concentration front arriving considerably before the peak concentration.

Formations that may prove to be particularly troublesome include the following.

1. Aquifers that contain narrow sections of extremely high permeability.
2. Aquifers with strongly anisotropic permeability conditions where the direction of flow will be inclined away from the hydraulic gradient and in the direction of maximum permeability.
3. Aquifers with geological properties that produce large-scale dispersion where contaminated water will flow in a broad band and exhibit a considerable spread between the arrival time of the concentration front and the peak concentration.
4. Aquifers that contain inhomogeneous* or heterogeneous sections, which result in anomalous groundwater patterns and rates of flow.

The next section includes some examples of formations that exhibit one or more of these conditions.

DARCY'S LAW

The water in an aquifer is in a continuous state of flow due to differences in hydrostatic head in different parts of the aquifer. In a homogeneous, isotropic aquifer this behavior can be described by Darcy's Law[31] in the form

$$V_w \approx 1.2 \times 10^{-3} \, I \frac{K}{p}$$

where

- V_w is the rate of flow of the ground water (mi/yr),
- K is the permeability of the aquifer, i.e., the number of gallons of water that can be transmitted through each square foot of aquifer per day per unit gradient (gal/day/ft^2) (typical values of K are listed in Table 5),
- p is the porosity, or the void fraction of the aquifer (typical values of p are listed in Table 6), and
- I is the hydraulic gradient, i.e., the difference in hydraulic

*As used in this report, "inhomogeneous" refers to a more or less continuous variation in permeability; "heterogeneous" refers to a discontinuous change in permeability.

head between two points divided by the distance between them.

A graphical representation of this equation using typical values of K, p, and I indicates that the rate of groundwater flow in any formation is determined predominantly by aquifer permeability (Fig. 3). Furthermore, both the porosity and the hydraulic gradient can be fairly well determined by standard techniques.[33,35]

Consequently, the succeeding sections of this part of the report discuss aquifer permeability, its measurement in the field, and the reliability of predictions of tritium transport based on this value.

AQUIFER PERMEABILITY

The permeability of a water-bearing formation is a measure of the rate at which it can transmit water under a given pressure. Laboratory determinations of its value can be made either directly, by observing the rate of flow through a sample of the formation under the influence of a known hydraulic gradient, or indirectly, by analysis of the size, shape, and arrangement of the grains comprising the sample.[34,36] Even when accurate values are obtained from laboratory samples, however, they represent the properties of only a small portion of the formation under consideration. Therefore, predictions of tritium travel over any appreciable distance still require field determinations of permeability.

Field values of permeability can be determined by any one of a variety of methods, all of which yield roughly the same value, but the method used most frequently is some variant of a pumping test.[33,36,37] Implicit in the interpretation of all such tests are two assumptions: (1) that the formation being tested is both isotropic and homogeneous, and (2) that groundwater flow is determined by the average hydrologic properties of this idealized formation. Thus, the permeability value determined by such a test is an overall average value that represents an integration of the effects of a variation with depth and direction as well as the effect of inhomogeneous

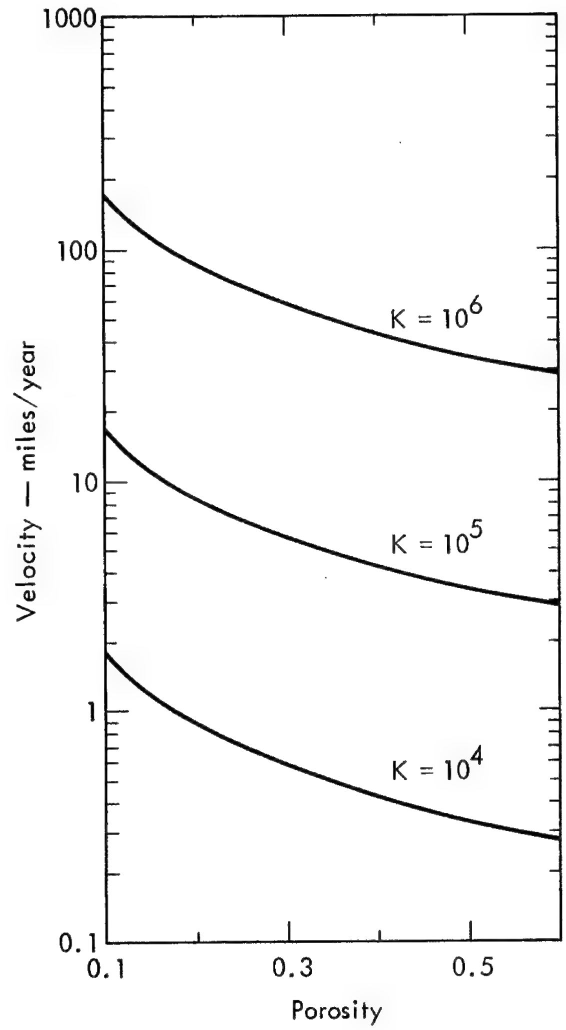

Fig. 3. Groundwater velocity as a function of aquifer permeability and porosity, for a 10/ft/mi gradient.

Table 5. Permeability coefficients K of typical geological formations (adapted from Ref. 32).

Formation	K mi/yr (unit gradient)[a]
Clays, loams, shales, solid igneous rocks, marbles	0 - 0.001
Cracked shales, recent loams, slightly cracked igneous rocks	0.001 - 0.01
Muds and very fine loamy sands, cracked rocks	0.01 - 0.1
Fine silty sands, old cracked rocks	0.1 - 1
Fine and middle, slightly loamy sands, fine sands, middle sands, slightly cemented sandstones	1 - 10
Coarse, uniform and slightly loamy sands, fissured rocks	10 - 100
Coarse and uniform sands, fissured rocks	10 - 100
Coarse and uniform sands, fissured limestones and marls	100 - 200
Well-sorted sands with gravel, fissured limestones	200 - 500
Fine and slightly sandy gravels, fissured and karstic cavernous limestones	500 - 1000
Medium and coarse uniform gravels, cavernous limestones	> 1000

[a] 1 gal/day/ft^2 = 9.26×10^{-3} mi/yr, or approximately 10^{-2} mi/yr.

and/or heterogeneous sections. However, the actual flow of groundwater through a particular section of an aquifer takes place in response to the hydrologic properties of that section. Therefore, predictions of flow based on field values may be highly misleading with regard to the actual flow of a significant amount of water that passes through a section of the aquifer whose properties vary appreciably from the overall or average properties of the aquifer. Particularly, difficulty will be experienced in predicting groundwater flow through formations in which (1) permeability varies with depth, especially in an aquifer containing a narrow section of extremely high permeability compared to the profile average; (2) permeability varies significantly with direction; (3) the geologic conditions result in large-scale dispersion; (4) geologic conditions such as channels, conduits, and faults are

Table 6. Porosity ranges of various materials.[31,33,34]

Material	Probable range (%)
Soils	50-60
Clay	45-55
Silt	40-50
Medium to coarse mixed sand	30-40
Uniform sand	30-40
Fine to medium mixed sand	30-35
Gravel	30-40
Gravel and sand	20-35
Sandstone	10-20
Shale	1-10
Limestone	1-10
Granite	1-10
Ringold	10-20
Fluviatile	20-30
Alluvium	30-40

present that result in anomalous flow rates and patterns. An example of each of the four kinds of formation is presented below.

Aquifers Containing Sections of Extremely High Permeability

In an aquifer of mixed coarse and fine materials (e.g., gravel interbedded with sands and silts), groundwater flows most rapidly through the coarse material by passing and flowing around water passing through the finer materials. Thus, almost all of the water that flows through an aquifer conceivably could pass through a narrow but highly permeable section of the aquifer at a rate many times the profile average. Well-pump tests of the San Joaquin Valley aquifer, which is composed of mixed sands, gravels, and clays and has a porosity of about 30%, indicated an overall average permeability of 100 gal/day/ft^2 (equivalent to a groundwater velocity of 0.1 mi/yr under the associated hydraulic gradient), whereas later measurements based on bomb tritium as a tracer indicated that almost all of the water flows through a narrow section of coarse gravel at a rate of 150 times this or at a velocity approximately 15 mi/yr.[38] A simple two-layer model of aquifer structure that agrees with both sets of measurements and the total reported thickness of the formation is shown in Fig. 4. Many other combinations of sectional thickness and clay and gravel type permeabilities, of course, would also satisfy the observed data. However, the significant point is that in almost all of these cases a simple model will show that 80 to 90% of all the water that passes through the aquifer could flow through a narrow section of the profile at a rate

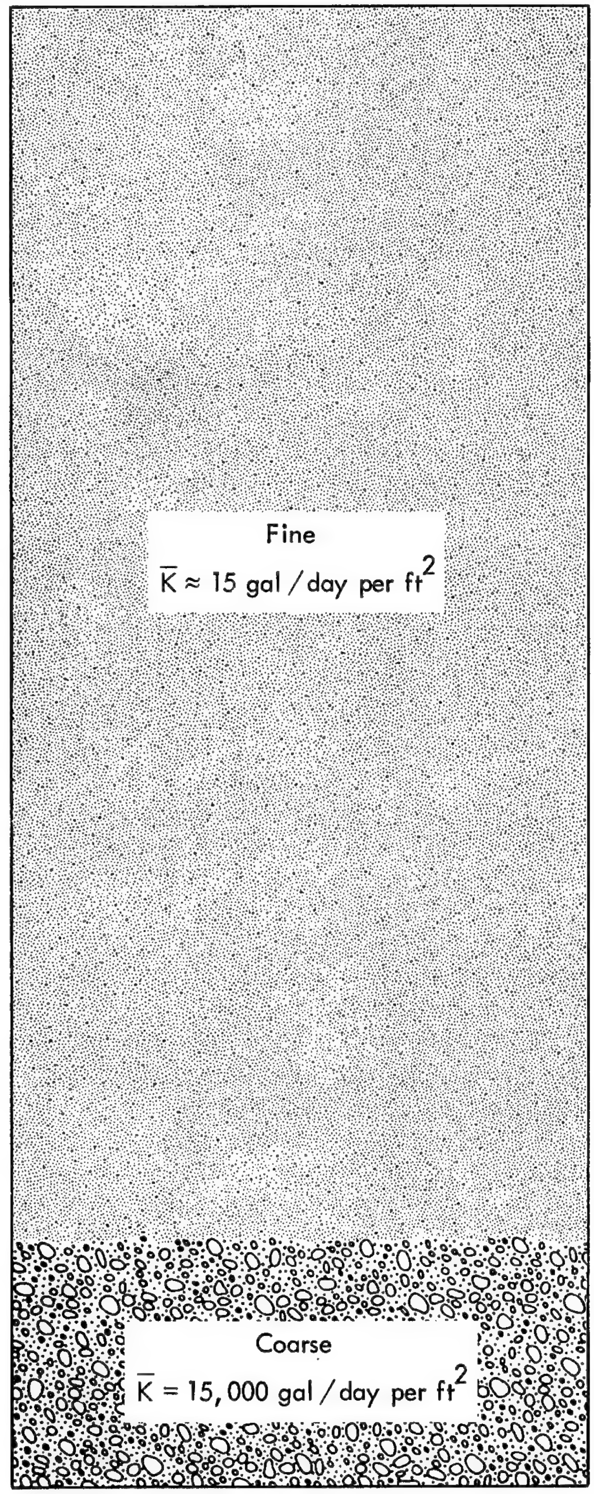

Fig. 4. Two-layer model of the structure of the San Joaquin Valley aquifer.

100 to 150 times that predicted on the basis of well-test permeability data for the formation.

Aquifers with Strongly Anisotropic Permeability

Geological formations are commonly structured so that the permeability in one direction is much greater than in another.[36] One consequence of this anisotropy is that the direction of flow is not in the direction of the maximum potential gradient, as is often assumed, but rather toward the direction of maximum permeability.[39]

Such anisotropic conditions can persist over long sections of an aquifer and thus cause the entire pulse to move in a direction significantly different than that of the hydraulic gradient. In fact, experiments on a small plot indicate that in a sufficiently anisotropic formation there may even be some contamination of the aquifer in an upgradient direction.[40,41] Furthermore, such conditions combined with the other possible inhomogeneities in an aquifer tend to intensify dispersive effects like those discussed below.

Aquifers That Exhibit Large-Scale Dispersive Effects

When a tritium pulse in the groundwater is transported through an aquifer, the center of the pulse travels with the average velocity of the groundwater. However, the individual units of tritiated water that comprise the pulse possess a wide range of velocities due to the random nature of their paths through the formation. As a result, an increasingly diffuse interface develops between the center of the pulse and the water it displaces.[42] This phenomenon is known as dispersion.

In isotropic and homogeneous formations, dispersive effects tend to be quite small, of the order of a few feet per mile. In nonideal aquifers, however, dispersive effects can be as much as two to three thousand times as high, as a result of anisotropic, heterogeneous, and/or inhomogeneous conditions. In sedimentary formations, for example, coarse lenses of greater permeability tend to draw the pulse out into a longer and thinner mass.[43] Similar effects can occur in basalt* and limestone formations as a result of solution cavities, lavatubes, and/or large variations in permeability.

One of the best examples of large-scale dispersion was observed in uranine dye tests at the Waste Disposal Site at Hanford, Washington. In the experiment, a concentrated dye was slowly introduced into the formation over a 24-hr period. Measurable concentrations of dye appeared in a down-gradient observation well (about 2 mi SSE of the disposal well) at about 67 days after injection and continued to appear until 227 days after injection; the peak concentration was observed at 129 days after injection.[37,44]

If the velocity is assumed to remain the same down-gradient beyond the observation well, then the front would have traveled some 2 mi farther by the time

*An example of observed dispersion in a basalt aquifer due to inhomogeneities in the formation is discussed in the section on Aquifers That Exhibit Inhomogeneous, Heterogeneous, and/or Other Geologic Conditions That Produce Anomalous Results (p. 17).

the peak concentration arrived at the observation well. The lateral spread of the dye while traveling this distance can be estimated from the finding that measurable concentrations of dye also appeared in a well about 2-1/2 mi from the injection site and about 30 deg east of the down-gradient observation well at about 67 days after injection. Thus, the observed lateral spread was about 1 mi for each 2 mi of down gradient travel. These data suggest that if an observation well were located several miles down-gradient from the disposal well in an extensive formation of this type, (1) contaminated water would begin to arrive at the observation well much sooner than predicted on the basis of average measurements, (2) the lateral spread due to dispersion which would develop over this distance would contaminate about half the total cross-sectional area of the aquifer.

Aquifers That Exhibit Inhomogeneous, Heterogeneous, and/or Other Geologic Conditions that Produce Anomalous Results

The National Reactor Test Site (NRTS) in Idaho covers an area of about 900 mi^2 in the north central part of the Eastern Snake River Plain. The principal underlying aquifer is a thick sequence of interlocking basaltic lava flows, highly porous and permeable in places, interbedded with varying thicknesses of a fine-grained completely impermeable sedimentary material.[44] The combination of this interbedding together with the essential anisotropy of the basalt system gives rise to a complex hydrologic system in which the movement of water is very rapid, erratic, and unpredictable, as is illustrated by the history of movement of radioactive wastes deposited in the disposal area.[45]

Disposals of radioactive wastes to the regional groundwater system have been made routinely, since 1953, through the disposal well of the ICPP (Idaho Chemical Processing Plant), which penetrates 150 ft below the water table. By 1960, 22 wells had been drilled in the area of the ICPP disposal well to an average depth of 630 ft below land surface. Because of the complexity of the aquifer system underlying the ICPP area, however, some of these wells tapped different aquifers (see Fig. 5). Consequently, although tritium was discovered on the NRTS site in August of 1960, the principal (D) aquifer underlying the ICPP was not defined until a year later, at which time all the wells within 3500 ft of the disposal well were redrilled to tap this aquifer (see Fig. 6). Thus, the first mapping of tritium concentrations in wells tapping the D aquifer did not take place until August 1961. The mapping indicated the presence of an anomalously high concentration of 850 pCi/ml at well 59 some 2300 ft to the Southeast of the disposal well (Fig. 6) where it would not be expected, because (1) the general direction of the groundwater gradient is toward the Southwest at 5 ft/mi, (2) it is almost three times the concentration at well 57, which is both down-gradient and in the direction of greatest permeability, and (3) it is more than three times the concentration at well number 51, which is approximately the same distance and travel time (see Table 7) from the disposal well.

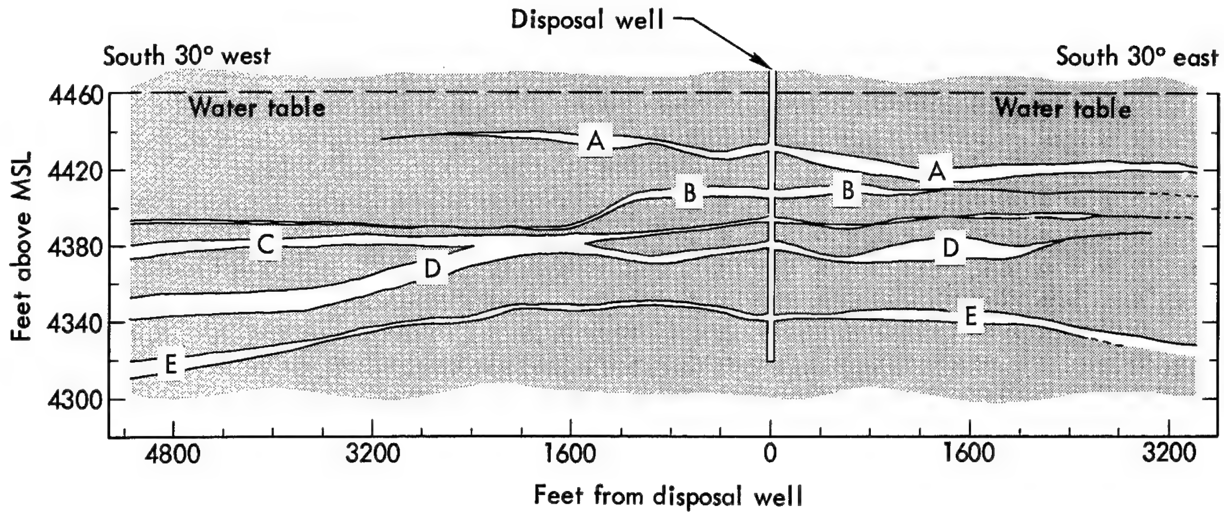

Fig. 5. Structures and thickness of aquifers A through E in the ICPP area along lines 30 deg West and 30 deg East from the disposal well.[44]

Table 7. Rate of flow from the ICPP disposal well to various observation wells, based on the disposal of 203 curies on December 10, 1961.[44]

Well No.	Distance from disposal well (ft)	Time of first arrival (days)	Rate of travel (ft/day)
47	703	5	141
43	850	6	141
41	925	11	84
46	1158	33	35
42	1336	34	41
52	1339	43	31
49	1347	55	24
48	1400	54	26
45	1625	39	42
59	2335	122	19
51	2354	103	23
67	3492	—	—
57	3502	80	44

As a result of a second mapping (see Table 7), the following observations can be made:

1. Wells about 1000 ft from the disposal well have flow rates of about 100 to 150 ft/day.
2. Wells about 1500 to 2500 ft from the disposal well have flow rates of about 20 ft/day.
3. The rate of flow to well 57 (3500 ft southwest of the disposal well) is 44 ft/day or 3 mi/yr.

4. At the time of the mapping, no measurable amount of tritium had reached well number 67 about 3500 ft from the disposal well.

Finally, the tritium disposal mapping of December 1962 (Fig. 7) shows the following:

1. The tritium-contaminated water has spread over an area encompassing 90 deg in traveling 4 to 6 mi from the disposal well.
2. An anomalously high concentration of 500 pC/ml was observed at well 38, which is 6000 ft from the disposal well.

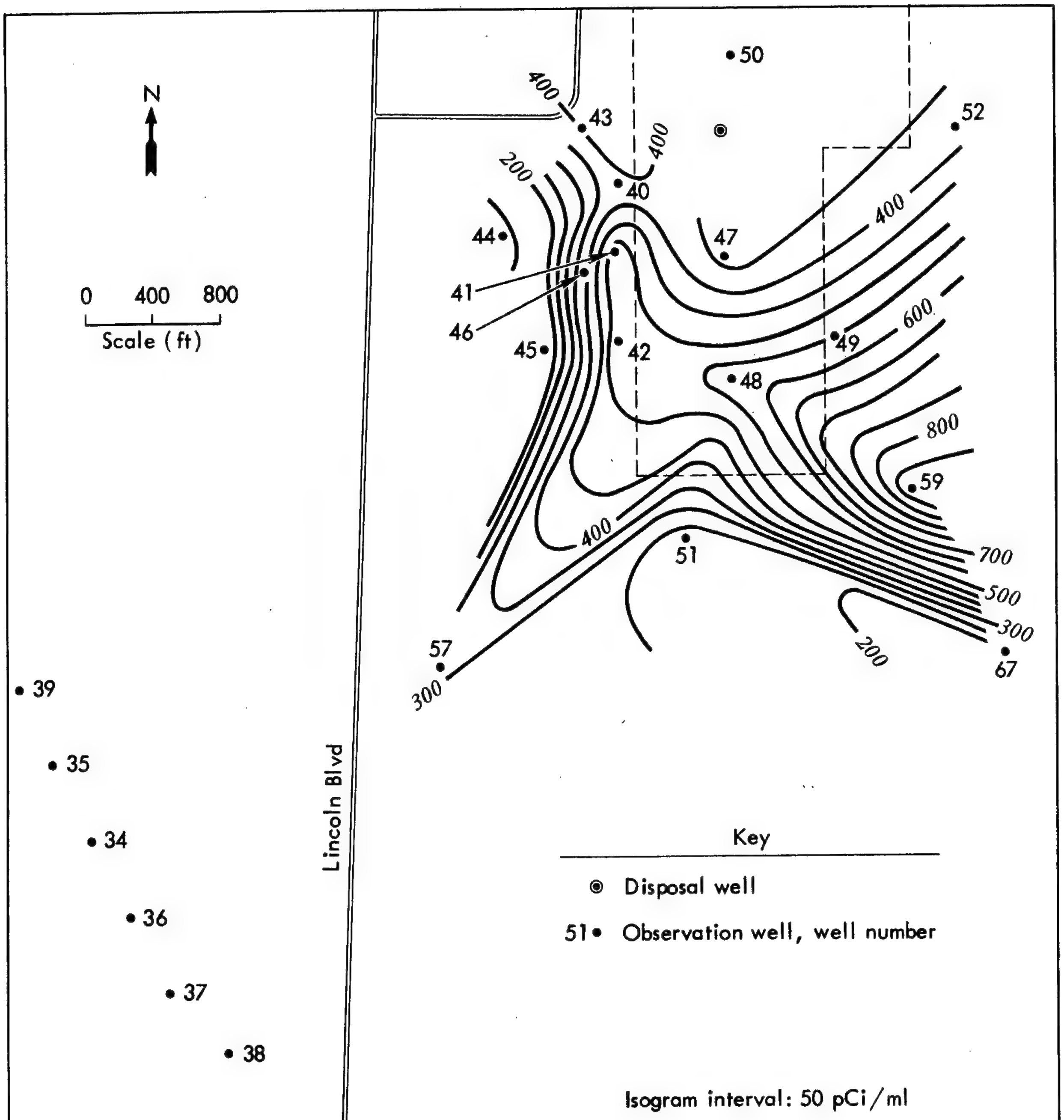

Fig. 6. The ICPP area showing tritium content (pCi/ml) of the wells that tapped the principal (D) aquifer on August 14, 1961.[44]

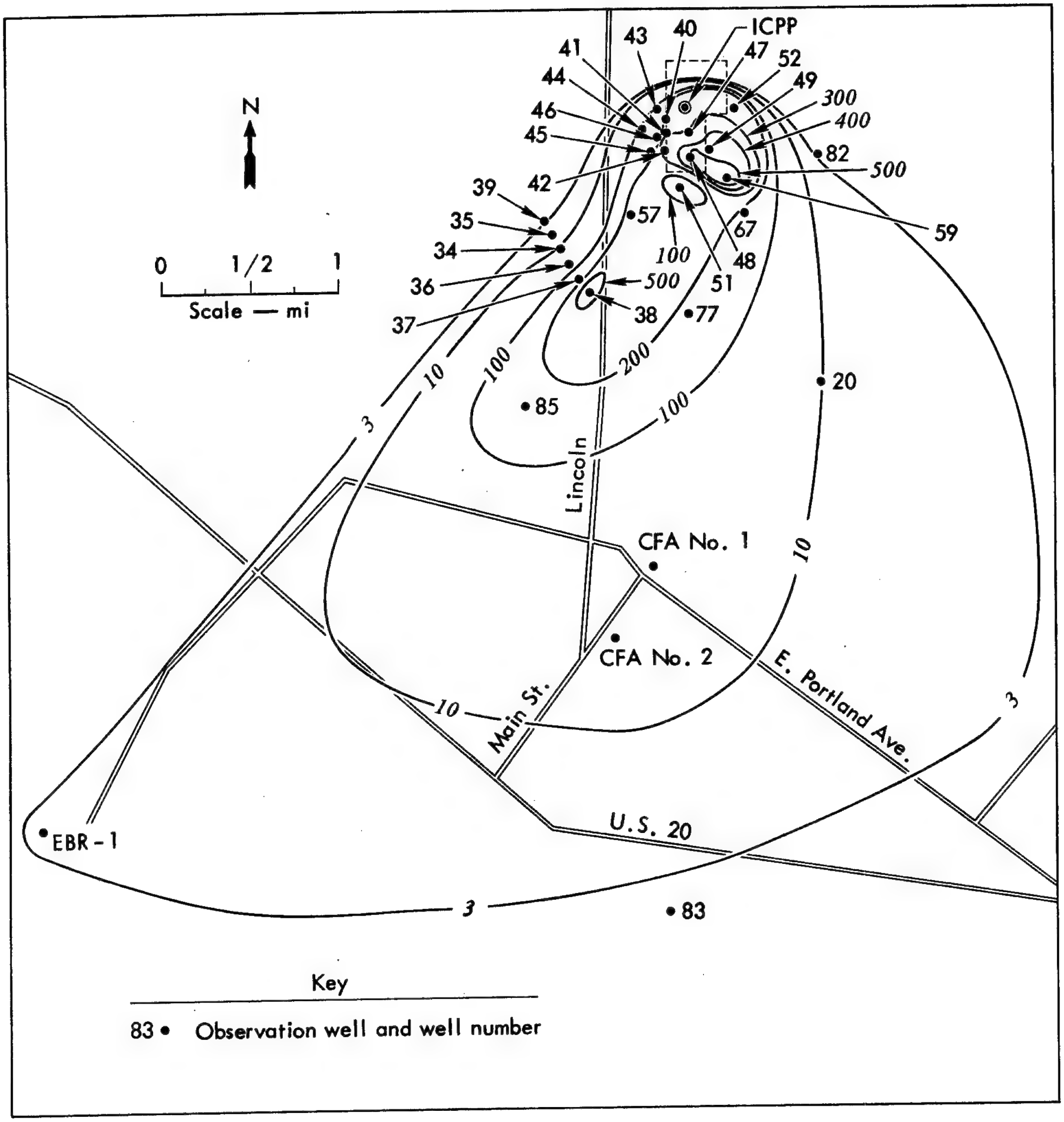

Fig. 7. The ICPP-CFA area showing the distribution of tritium in the regional groundwater, December 1962.[44]

These results imply that in a formation as complex as this, virtually nothing can be predicted, on the basis of general knowledge of conditions at the site, about the flow pathways and travel times of groundwater even over short distances, to say nothing of the long distances that may be of interest in cratering situations. In fact, considering the difficulty of locating the principal aquifer and the various anomalies found by subsequent mappings, even a very sophisticated and detailed network of well and tracer tests in one

section of the formation gives very little information about conditions pertaining in other sections of the formation. Finally, significant lateral spread is observed once again, indicating that although the waste has traveled only about 5 miles down-gradient from the disposal well, an area of about 25 mi^2 is contaminated.

CONCLUSIONS

The tritium pulse produced by a large-scale nuclear cratering detonation can be transported to a point of potential water use by flowing groundwater. To a first approximation, this process can be described by Darcy's Law. Predictions of the travel time and the associated decrease in activity between the two points in question can therefore be made once the geohydrologic properties of the area involved are known. However, almost all methods of field measurement of aquifer properties are based on two assumptions: (1) the aquifer is homogeneous and isotropic, and (2) groundwater flow is determined by the average hydrologic properties of this idealized formation. Thus the values determined are overall average values that represent integrations of the effects of the variation of aquifer properties with depth and direction, as well as the effects of inhomogeneous and/or heterogeneous conditions. Consequently, in certain geologic situations, flow calculations based on field values of aquifer properties may be highly misleading if they are assumed to predict the actual behavior of a significant fraction of the total amount of water contaminated by the pulse. Examples of the types of formation in which this problem is likely to occur are as follows:

1. Aquifers containing narrow sections of extremely high permeability, illustrated by the San Joaquin, California aquifer where tritium tracer measurements indicated that between 80 and 90% of all the water passing through the aquifer flows through a narrow section of coarse gravel at a rate 150 times that calculated for the aquifer by well-tests.
2. Aquifers with geologic properties, including large-scale anisotropies, that produce large-scale dispersion, illustrated by the aquifer at the waste disposal site at Hanford, Washington, where a dye solution dispersed in such a manner that over the course of a 2-mi path, (1) it developed an angular spread of 30 deg, and (2) the concentration front took only half as long as the peak concentration to reach the observation well.
3. Aquifers that contain inhomogeneous and/or heterogeneous sections that result in anomalous groundwater flow, illustrated by the aquifer at the Idaho National Reactor Test Site, where tritium migration mappings revealed (1) high rates of flow (140 ft/day) near the disposal well, (2) an anomalously high concentration of tritium in a well about 4000 ft from the disposal well, and (3) the development of an angular

spread of 90 deg over a path length of 4 to 6 mi.

Finally, unless measurements of aquifer properties, especially permeability, are sufficiently detailed to resolve the effects of variation with depth and direction as well as the effects of inhomogeneous and/or heterogeneous aquifer structure, tritium-contaminated water either (1) can arrive where it is expected much sooner than expected, perhaps by a factor of 100, (2) can arrive where it is not expected at all, or (3) can spread and travel in a much broader band than expected, with the concentration front arriving considerably earlier than the peak concentration. These conclusions clearly indicate that accurate preshot predictions of the hazard to man from contamination of the groundwater environment as the result of a nuclear excavation project requires detailed and extensive study of the aquifer involved. In any specific situation, the amount of information needed for this task will be a function of the complexity and heterogeneity of the system in question. Furthermore, in many real situations, even after detailed and extensive preshot study of the aquifer, we can be confident that tolerance levels are not being exceeded only after extensive postshot monitoring.

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